Kinematic interpretation of the 3D shapes of metamorphic core complexes

Laetitia Le Pourhiet, Benjamin Huet, Dave May, Loic Labrousse, Laurent Jolivet

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Authors:

Laetitia Le Pourhiet [1,2]

Benjamin Huet [3]

Dave A. May [4]

Loic Labrousse [1,2]

Laurent Jolivet [5]

Affiliations:

1 UPMC Univ Paris 06, UMR 7193, ISTEP, F-75005, Paris, France.

2 CNRS, UMR 7193, ISTEP, F-75005, Paris, France

3 Department for Geodynamics and Sedimentology, University of Vienna, Althanstrasse 14 A-1090 Vienna

4 Institute of Geophysics, Department of Earth Sciences, ETH Zurich, Switzerland

5 ISTO, Université d'Orléans-CNRS, UMR 6113, F-45071 Orléans, France

Abstract

Metamorphic Core Complexes form dome shaped structures in which the ductile crust is exhumed beneath a detachment fault. The 3D dome geometry, inferred by mapping the schistosity in the exhumed crust, can be either elongated normal to the stretching direction or along it. In the first case, the domes are interpreted as having formed during extension. However, in the second case, they are interpreted either as strike-slip, transpressive or constrictive extensional structures, depending on the geodynamic context. Numerical models of metamorphic core complexes published to date are all two-dimensional and therefore, theoretically only apply to domes which are elongated normal to the stretching direction. Here, we explore by means of 3D thermo-mechanical modeling, the impact of 3D kinematic extensional boundary conditions on the shape of metamorphic core complexes. We examine the impact of a transtensional step over and of horsetail splay fault kinematics on the dynamics of exhumation, finite strain and P-T paths, and compare them to cylindrical 3D models. We show, for the first time, that domes formed in transtensional step over, or at the tip of propagating strike-slip faults, display a finite strain field which can be interpreted as characteristic of a transpressive domes, although no shortening was applied in the far-field. Applying our models to the Cyclades, we propose that the coeval formation of domes elongated normal and parallel to the stretching during the Miocene can be the result of horsetail splay fault kinematics, which could correspond to the formation of a tear in the Aegean slab.

1 Introduction

Metamorphic Core Complexes (MCCs) form dome shaped structures of metamorphic rocks exhumed from the middle to lower crustal depth under a detachment fault. These domes are present in many post-orogenic contexts. The dome axis, which is recognized by mapping the schistosity pattern in the exhumed metamorphic rocks, can either be elongated parallel to the stretching lineation (a-type domes Figure 1, orange structure), or normal to this direction (b-type domes Figure 1, blue structure). In the first case, the schistosity strikes mainly parallel to the direction of stretching while in the second case it strikes normal to it. Hence, b-type domes are cylindrical structures and may be modeled in 2D plane strain cross-section while the a-type domes are intrinsically 3D structures since local horizontal constriction occurs normal to stretching. This a-type versus b-type classification, first defined in Jolivet et al. [2004] based on Mediterranean examples, relies on a quantitative description of the finite-strain field. The interpretation of these structures in terms of larger scale kinematics is an ill posed inverse problem, which reaches by definition, a non-unique answer.

Structural geologists generally unambiguously interpret b-type domes as the mark of extension.

This interpretation could be questioned, however, since the structure is cylindrical, it constitutes the simplest conceptual model. The Bitterroot Eocene core complex in the Northern Rockies, Idaho [*Hyndman, 1980*] and the North Cyclades Detachment System (NCDS), Greece [*Jolivet et al., 2010*] constitutes a set of excellent natural examples of b-type structures, although in both cases, strike-slip faults are inferred in the vicinity of these domes [*Foster et al. 2007; Philippon et al. 2012*].

However, the kinematic interpretation of a-type domes is not straightforward and depends on the geodynamic context. In some cases like Kesebir-Kardamos, Bulgaria *[Bonev et al., 2006*], Buckskin Rawhide, Harcuvar; Arizona, US [*Reynolds and Spencer,1985; Howard and John 1987*], Naxos, Greece [*Kruckenberg et al., 2011*] or the Western Gneiss Region, Norway [*Labrousse et al. 2004*], the domes are mainly associated to extension with a slight component of strike-slip in the kinematics. In other cases, these domes are principally associated with strike-slip kinematics with a component of transtension like the Dai Nui Con Voi dome, Vietnam, [*Jolivet et al., 2001*] or a component of transpression like along the Red River Fault system [*Leloup et al., 1995*]. However, in the Betics Cordillera [*Martinez-Martinez et al. 2002; Augier et al. 2005*] or in the Variscan orogeny (e.g. Montagne noire, France, [*Echtler et Malavielle, 1990*]), the large-scale kinematics that accompanies their formation is deeply debated. Whether these structures are interpreted as the result of shortening followed by perpendicular stretching, of coeval stretching and lateral shortening, or of pure strike-slip wrenching changes the interpretation of these structures in term of the large scale kinematics at the time of their formation.

Since Buck [*1991*], MCCs are known to form in a local extensional setting when the viscous diffusivity of the flow in the lower crust is small, i.e. when lower crust is thick and/or weak. Several processes leading to either, or both, of these conditions have been proposed and modeled: (i) thermal equilibration of a thickened crust at high Moho temperatures [*Gaudemer et al., 1988; Block and Royden, 1990; Tirel et al., 2008; Tirel et al., 2009*], (ii) partial melting of the lower crust [*Lister and Baldwin, 1993; Vanderhaeghe and Teyssier, 2001; Rey et al., 2009*],

(iii) adjunction of water [*McKenzie and Jackson 2002*] or (iv) underthrusting of weaker units below stronger ones prior to extension [*Huet et al. 2011a, 2011b*]. All the models stated above only accounted for cylindrical structures and therefore are theoretically only applicable to b-type domes.

Here, we first attempt to understand the impact of 3D kinematic extensional boundary conditions on the 3D geometry of MCCs, i.e. the relative geometry of the schistosity pattern versus the stretching lineation pattern within the exhumed crust. In the second part of the paper, we compare the results of the model to the MCCs of the Cyclades, which provides examples of the two types of domes formed during the same, well constrained, geodynamic event. Lastly we draw more general conclusions about the kinematic implications of the observation of a-type domes.

2 Modeling approach

2.1 Choice of initial conditions

The models have a fixed sized geometry of 200 x 200 x 100 km and evolve in time for 12 Myr, with a constant extension rate of 1cm/yr. Given the assumed size and the grid resolution (3 km) of the models, we expect to resolve the first order impact of large scale, 3D kinematic constraints. The models consist of three horizontal layers (See Figure 2 for the vertical profile).

Since the thermo-mechanical initial conditions for forming MCCs are well understood from 2D modeling, and our study focus on the impact of 3D boundary conditions on their resulting shape, the model design accounts for most of the factors that are known to favor the occurrence of MCCs. This includes an initially thickened crust of 50km [*Buck 1991, Gaudemer et al. 1988, Block and Royden, 1990*] and an initial thermal gradient which is set to 17.5°C/km.

We note that this gradient yields a Moho temperature of 875°C, which is higher than the 800°C limit proposed by Tirel et al. [*2008*]. At asthenospheric depth, the initial temperature is clamped not to exceed 1300°C (Figure 2). The crust itself consists of two layers of 25 km each. At a given temperature, the top layer (brown shades, Figure 2) is mechanically stronger than the lower one (blue shades, Figure 2), imposing a step in the strength profile at the interface between the two layers. These models thus resemble the set up of the pioneering analogue models of MCCs [*Brun et al. 1994*], except they account for the re-strengthening of material as it gets exhumed and cools.

The initial weakness of the lower layer can be equally interpreted as the result of adjunction of water in the lower crust [*McKenzie and Jackson 2002*], as the occurrence of partial melt in the lower crust [*Lister and Baldwin, 1993; Rey et al., 2009*] or as the presence of under-plated soft sediments or continental upper crust [*Huet et al. 2011a*]. We do not dismiss that the formation of MCCs can be of diapiric nature [*Lister and Baldwin, 1993*], however, in order to keep the model general, we assign a constant density to the whole crust of 2800 kg.m^{-3} so that buoyancy forces are limited to those arising from the thermal expansion coefficient which is set constant to $3x10^{-5}$ K^{-1}.

2.2 Treatment of the rheology

At the scale of these models, the rheology in any given volume has no reason to be one of pure quartz, pure olivine or plagioclase and must be influenced by the layering of the rocks, their anisotropy, structural softening and hardening due to small scale boudinage and folding. In practice it is not yet possible to account for all this complexity in crustal or lithospheric scale modeling. As the viscous strength of rock depends to first order on temperature, we approximate it by a Newtonian flow rule based on Frank-Kamenetskii, for which we need to provide two parameters, a reference viscosity η_0 and a characteristic temperature θ^{-1}, in order to compute the effective viscosity,

$$\eta_{eff} = \eta_0 e^{-\theta T} \tag{1}$$

At high stress, this viscosity is lowered by the action of a Drucker Prager visco-plastic flow rule such that

$$\eta_{eff} = \frac{1}{2}\frac{\sigma^y}{\dot{\varepsilon}^{II}}, \tag{2}$$

where $\dot{\varepsilon}^{II} = 1/2\sqrt{\dot{\varepsilon}_{ij}\dot{\varepsilon}_{ij}}$ is the second invariant of the strain rate tensor ($\dot{\varepsilon}_{ij}$) and the yield stress, σ^y, depends on pressure P and plastic strain ε^p accumulated by the particles following

$$\sigma^y = P\sin\phi(\varepsilon^p) + C_0\cos\phi(\varepsilon^p). \tag{3}$$

Accounting for plastic strain softening is necessary [*Lemiale et al. 2008*] to localize plastic shear bands in visco-plastic codes such as GALE [*Moresi et al 2003*]. We impose softening on friction in the upper and lower crust with a friction angle decreasing from 30° to 10° with plastic strain varying from 0 to 20%, given the element size (3km), softening is therefore achieved for a displacement of 500 m in nature.

The rheological parameters used for the study are listed in Table 1 and were chosen because they resulted in a yield strength envelope similar to pre-existing 2D models of Huet et al. [*2011a*], thereby allowing us to verify (or benchmark) our results with pre-existing models run with a other numerical code, namely FLAMAR [*Yamato et al. 2007*]. In the parameterization, we have chosen to keep η_0 constant and to vary the characteristic temperature $1/\theta$. This choice was governed by the idea that when the temperature is zero at the surface, both the lower and upper crust must have the same effective viscosity in order to enable yielding. However, with this choice, the viscosity of the lower crust may drop to a non-realistic value at the base of the lower crust. Therefore we adopted a lower cut-off for the viscosity in the models, which we set to 10^{19} Pa.s, as in previously published 2D models [*Huet et al. 2011a, Tirel et al. 2008*].

The previous study by Huet et al. [*2011a*] investigated the role of the rheological layering and temperature on the distribution, or localization of the deformation without imposing a priori the location of faults. To avoid strain localization on the side of the box, they used random damage all through the upper crust except in the vicinity 30 km from the boundary. The model domain was longer and show that with a 25 km thick brittle crust, strain started to localize by distributed necking with a wavelength of 70 km. Depending on the rheological layering of the crust, either the strain remained distributed and no exhumation of lower crust was occurring or only one of the initial grabens was still active after 2 Myr before turning into a MCC. In the present study, we investigate the role of 3D kinematics on the shape of MCC. We have therefore chosen a rheological layering which favor their formation and imposed a 50 km wide zone with random pre-existing damage in the center of the model to avoid unwanted border effects.

2.3 Numerics

All the numerical experiments were run with GALE 1.6.1, an open source code, which solves incompressible Stokes flow coupled with heat conservation via a finite element, particle-in-cell method in three dimensions [*Moresi et al., 2003*]. The computational mesh consisted of 64x64x32 Q1 (trilinear) elements. Evolution in time is obtained through advection of particles (tracking lithology) and the thermal energy equation, with time steps limited to 10% of Courant criterion.

As GALE uses Q1Q1 elements, it is necessary to scale the viscosity of the model to be close to one in order to obtain accurate results and convergence. The scaling used to run the code is provided in Table 2. Additional passive markers have been included to compute the finite strain field in the lower crust and to track P-T paths. An Octave script used for post-processing the finite strain ellipsoid and exporting it into a VTK format for visualization purpose are included as supplementary material. The method used to compute the finite strain tensor in 3D from passive markers is detailed in appendix A.

Each run consists of 200-300 time steps, requiring approximately 72 hr of computation using a parallel geometric multi-grid (GMG) solver on 16 CPU (dual core AMD Opteron 2.6 GHz, each with 1 GB of RAM). A full description of the multi-grid method used in GALE is provided in Appendix B. We refer the reader, who would like to reproduce the results or reuse the input files, to the supplementary material. It includes the input files of the models, solver options and patches for GALE to reproduce the specific boundary conditions and viscous cutoff implemented for this study.

3 Impact of boundary conditions on the shape of MCCs

Three types of extensional boundary conditions are considered. The boundary conditions are represented by black arrows that indicate the velocity applied on the front (pink) and back (cyan) boundaries of the models (Figure 3a). First, 3D cylindrical extension is applied by imposing a constant normal velocity on one side of the model (the cyan boundary, Figure 3a and 4). This model is used both for verifying the modeling approach as compared to existing 2D models and to compare the results with a non-cylindrical, 3D boundary conditions. The second model considers extension occurring at an extensional step-over between two left lateral strike-slip faults applied on the front (pink) and back (cyan) boundaries of the model (Figure 3b and 5). The third model considers the case of a transtensional fault propagator which accommodates a left-lateral step in the rate of extension at the front of the model (pink) (Figure 3c and 6). The two lateral boundaries are free-slip boundaries.

3.1 Surface deformation

The surface deformation pattern after 12 Myr of extension is outlined in map view, by the light brown stripes, which were originally forming square (Figure 3). The most striking feature at that point is that the cylindrical and the fault propagator kinematics lead to the exhumation of deep crustal material to the surface (blue material on Figure 3a, 3c), whilst the step-over kinematics model does not (Figure 3b). In the case of cylindrical boundary conditions (Figure 3a), the model produces one structure, elongated normal to the direction of stretching. In the case of fault propagator kinematics, the initially deep crustal material is exhumed along a curvilinear trend which follows the fault propagator and turns to become orthogonal to the free slip side of the model. This trend is therefore parallel to the stretching direction close to the back boundary, whereas it is perpendicular to the stretching direction close to the left hand side boundary. Looking in more detail, one sees that in between the two branches, there are less exhumed rocks and that the strike-slip part appears to be elongated further towards the back of the model than the location of the cylindrical part of the exhumed structure (Figure 3c).

3.2 Topography

Colored isolines indicate the surface topography (Figure 3). All the models produce relatively low topographic expression with an altitude ranging from -800 m to 800 m for the model 1 (Figure 3a), from -1200m to 1000m for model 2 (Figure 3b) and from -1200 m to 1300 m for model 3 (Figure 3c). The exhumed deeper crustal rocks tend to be located below topographic lows rather than beneath the topographic highs. Noticeably, the two models with a strike-slip component produce significantly more topographic expression than the cylindrical model. The steepest gradients are observed along the strike-slip faults. Along the front side, the topography may be over estimated due to border effects, but in the center of the model, the topographic gradients remains significant. In the regions exhibiting strike-slip deformation, the topographic lows tend to be aligned in the direction of stretching. Comparing the cylindrical model (Figure 3a) with the cylindrical part of the fault propagator model (Figure 3c), the presence of the strike-slip fault changes the geometry of topography, since a secondary topographic low that is absent from the purely cylindrical model appears.

3.3 3D-geometry and finite-strain inside the domes

Along cross-sections through all the models, one sees that the Moho remains strikingly flat (Figure 4, 5 and 6) while the deeper, ductile, part of the crust is exhumed within dome shaped structures, even in model 2 where it does not reach the surface. The finite strain field calculated locally in the lower crust is represented by the stretching lineation superposed to the geometry of the models (cylinders colored in function of the strike of the lineation). The lineation and the pole of the schistosity are also plotted on stereo-diagrams.

In the case of cylindrical extension (Figure 4), the dome is approximately symmetric and elongated normal to the stretching direction. The detachment fault at the top of the lower crust is outlined by nearly horizontal lineations parallel to the stretching direction (Figure 4a and red dots on Figure 4b), whilst vertical lineations delineate the deep core of the dome. The poles of the schistosity are aligned on a vertical great circle parallel to the stretching direction (black dots on Figure 4b). This pattern reflects the cylindrical nature of the dome. However, a few vertical schistosity planes that strike parallel to the stretching direction mark local constriction. The similarity between this cylindrical model and the 2D models on which we have based the model set up [*Huet et al., 2011a*] provides confirmation that although the models were run with a lower resolution in 3D, the first order kinematics is maintained. The dome formed in this simulation displays all the characteristics of a b-type dome.

The extensional step-over kinematics (Figure 5) leads to the localization of a dome in the center of the model. In this simulation, the deeper ductile part of the crust is exhumed with the shape of a dome beneath the topographic depression formed by normal and strike-slip faults in the upper brittle part of the crust. This dome is elongated in the direction of stretching imposed from the boundary condition. It is a non-cylindrical structure because the lineations (Figure 5a) are pointing out of plane when considering a cross section taken normal to the elongation of the dome. The lineations (red dots Figure 5b) within the dome are systematically horizontal and the schistosity planes (black dots Figure 5b) are approximately vertical, except at the very top of the dome forming an antiform structure. The particles show that this structure is slightly asymmetric with its steepest dipping limb located on the side of the strike-slip fault.

In this simulation, the strain is highly partitioned with depth. In the upper crust, the deformation takes place along steep grabens or pull apart basins (Figure 3b) which reflects the extensional nature of the boundary conditions. However, in the deeper part of the crust, the lineation forms a horizontal sigmoid compatible with pure left-lateral deformation, while the ductile exhumed lower crust within the dome deforms by constriction and forms an antiform (Figure 5a-b). The deformation in the exhumed part of the deeper crust is characteristic of an a-type dome (Figure 1) with a strong local component of constriction, which was not imposed in the boundary conditions (Figure 3b). Dynamically, the formation of this structure is due to the local depression that forms above the extensional step-over. The large topographic gradient drives the flow in the ductile crust, attracting it towards the pull apart, where it rises to form the dome. As the dome grows, the strength of the upper crust drops locally, leading to strong strain localization.

The fault propagator kinematics (Figure 6) display the formation of two independent domes linked by a transfer zone. The first dome forms along the gray boundary with an almost cylindrical shape, similar to the dome of Figure 4a, although the lineations at the front of the structure are slightly oblique to the stretching direction (indicated by yellow colored cylinders instead of gray cylinders). The second dome forms at the location of the imposed strike-slip boundary condition. As for the a-type dome of Figure 5a, the structure is an asymmetric fold with a steeper dipping limb close to the fault. However, the lineations in the deeper part of the crust (Figure 6a) tend to strike oblique to the direction of stretching by an angle of 25 to 50° (red dots on Figure 6b) and indicate that this dome has been fed from both the back (cyan) and the right (black) side of the model.

This model with mixed boundary condition shares many common characteristics with model 1 and 2. However, in this mixed mode, the orientation of the stretching lineation in the first 10 km of the crust is much more spread out than for purely a- or b-type domes. In this mixed mode, the azimuth of the lineation may locally rotate by 40-50° as compared to the direction of extension or shear imposed from the boundary conditions.

3.4 P-T paths

The computation of synthetic P-T paths is common practice to validate models with observations in orogenic settings [*Gerya et al. 2000, Yamato et al. 2007*]. In both the case of the cylindrical and the fault propagator models (Figure 4c, 6c), despite the rather different shape of the domes in map view and the different kinematic imposed at the boundary, the P-T paths recovered from the two models are very similar. In both cases, the rocks are exhumed from all depths along the initial thermal gradient of 17°/km and starting from there, the exhumation takes place along isothermal or slightly heating P-T paths. In both models the final cooling of the exhumed rocks follows a linear gradient of c.a. 50°C/km which reflects the thinning of the crust. P-T paths of the step-over model (Figure 4c) contrast with the first two. They do not display heating P-T paths, and the decompression arises with significant cooling. The thermal gradient after 12 Myr reaches only 33°/km as a result of the relatively small amount of thinning of the lower crust involved in these models as compared to model 1 and 3.

4 Summary and discussion of modeling results

4.1 Impact of initial conditions

An important feature of all these models is the rheological step that was introduced at mid crustal level. As discussed earlier, this step may represent a change in lithology due to nappe stacking as in Huet et al. [*2011a; 2011b*], or could be caused by the presence of fluid [*McKenzie and Jackson 2002*] or be the limit of partially molten crust [*Lister and Baldwin, 1993; Vanderhaeghe and Teyssier, 2001; Rey et al., 2009*]. This step in viscosity is important for the outcome of the models as it allows localizing the deformation on a single detachment zone as shown and discussed in detail in Huet et al. [*2011a*] and it facilitates the rise of ductile material beneath the pull-apart in model 2. Similarly, we did not explore the effect of the initial depth of the step in viscosity, but we contend it would affect the initial spacing of the faults and therefore potentially change the outcome of the models by increasing the number of domes if the step was located at a shallower depth [*Bullard, 1936; Vening Meinesz, 1950; Spadini et Podladchikov, 1996*]. If this step was to be at a lower depth, several domes would probably form initially, but at larger strain one would localize strain more efficiently [*Lavier et al., 2000*; *Wijns et al., 2005*]. If several steps were to be included, we can predict that the localization of the deformation would probably jump from the deepest one to the shallower one over time as was observed in Huet et al. [*2011a*]

4.2 Rate of extension versus rate of exhumation

We observe that the model with transtensional splays (model 3, Figure 6c) exhumes lower crustal rocks to the surface along warmer geothermal gradients compared to pull-apart models (model 2, Figure 5c) for the same rate of extension. Part of this discrepancy is due to a geometrical effect of the design of the model because the pull-apart modelled here is *infinite* and therefore the source of material is limited to the left and right side, whilst in the case of a splay fault, significant amount of material is pumped from the back side (cyan side, Figure 6) of the model. Moreover, as the crust thins much more in the case of a splay fault (Figure 6) than in the case of a pull-apart, and considering that the strength of the lower crust or the mantle is not sufficient to maintain steps in the Moho, the lower crustal material located on the side with no extension is forced to flow towards the a-type structures. The second reason for the discrepancy is that as the dome gets wider, the diffusivity of the flow entering the a-type part of the dome increases as the cube of the thickness of the channel [*Buck, 1991*], enhancing exhumation rates.

One may also question the effect of changing the rate of extension on the outcome of our models. Based on analogue models of sand and silicone putty, Brun et al. [*1994*] and Brun [*1999*] suggested that increasing the rate of extension decreases the rate of exhumation. This rate dependence is mainly due to viscous coupling which competes with the diapiric ascent of light silicone putty into the sand. As it was discussed in *Wijns et al.* [*2005*], the change of the exhumation rate observed in analogue models is largely reduced when the buoyancy forces are small, which is the case in the model presented in this paper where buoyancy forces are kept to minimal and most of the exhumation is kinematically driven. Therefore, if one would decrease the extensional velocity, the rate of exhumation would probably decrease (but not drastically), in response to both kinematic forcing and reduced buoyancy forces in response to cooling by diffusion. The P-T paths would reach a cooler final thermal gradient, but as the rheological step remains, the overall kinematics would not change drastically. Similarly, increasing the rate of extension would probably increase the rate of exhumation and render the P-T path warmer in response to increased buoyancy forces, as was observed in models including melts [*Rey et al., 2009*]. In any case, the structures and their orientations, which are central to our analysis, would remain unchanged. The presence of the rheological step thus render the overall model kinematics, or finite strain pattern, almost independent of the rate of extension.

4.3 Significance of a-type domes

We did not manage to create a-type domes with purely cylindrical boundary conditions, even though local components of constriction were observed in model 1. We note however that this could be an issue with the numerical grid resolution used in these models. It is possible that a-type structures could form at segments of the boundaries between domes if we considered similar instabilities as the one proposed by Gerya [*2010*] for the formation of transform faults at mid oceanic ridges. We can also posit that similar dome shapes would occur if we had imposed the formation of two offset domes in the initial conditions such as those used in [*Choi et al., 2008; Allken et al. 2011*].

In any case, this does not contradict the main conclusion of our modelling study: when a transtensional strike-slip component is involved in the deformation of hot post-orogenic (i.e. previously thickened) crust, a-type domes form. The most important result is that the constrictive ductile deformation within these domes is the mark of strike-slip transtensive faulting in the upper crust, rather than shortening normal to the stretching direction. We agree that at a first glance these structures resemble folds, however, the contact between the upper and the lower units is used as a detachment fault to exhume deeper units within the dome. Finally, the modelled topographic pattern indicates that sedimentary basin should form on the lateral side of a-type domes.

5 Comparison of the models to the Aegean MCCs

The Aegean Sea is a post-orogenic domain thinned in the back arc of the Hellenic subduction zone [*Le Pichon and Angelier, 1981*]. In the Cyclades, the two types of Metamorphic Core Complexes have been exhumed with Naxos and Tinos being considered as representative of a-type and b-type domes, respectively [*Jolivet et al., 2004*] (Figure 7a). In this section, we discuss the applicability of our models to this domain and the implications for our understanding of the Aegean dynamics.

The P-T paths obtained are not sufficiently different from the natural P-T paths of Naxos [*Martin 2004, Duchène et al. 2006*] or Tinos [*Parra et al., 2002*] to reject either model based solely on thermo-barometric validation. However, in the fault propagator model, the behavior of the two parts of the dome differs. Whilst rocks are exhumed from all depths close to the free-slip boundary (dark gray, Figure 3f), within the strike-slip part of the dome, only high temperature paths are sampled (light gray, Figure 3f).

The lineations measured in the MCCs of the Cyclades are reported on Figure 7a. The outline of the modeled area is projected on the map so that the strikes of the lineation along the North Cycladic Detachment System (green arrows, [*Jolivet et al., 2010*] are parallel to the direction of stretching of the models. In that frame, the roughly north-south lineations of Naxos, Paros and Ios (red arrows) [*Huet et al., 2009; Gautier et Brun, 1994*] form a 30-60° angle with the direction of stretching of the model and the late lineations along the South Cycladic Detachment System appear with intermediate strike (orange arrows, [*Isgleder et al., 2009*]). Not only is the geographic repartition of the lineation in good agreement with the fault propagator model (Figure 3c) but also the domes of Naxos, Paros and Ios fall actually in the class of domes that are elongated in the direction of the lineation. This observation, together with the warmer P-T paths in these domes and the migration to the south with time of from top-to-the-north-north-east detachments to a top-to-the-south detachment [*Grasemann et al., 2012*], leads us to conclude that the transtensional fault propagator model is a valid model for the formation of the Cyclades. The obliquity of the lineation in the central Cyclades was previously attributed to a late phase of solid rotation around a vertical axis [*Morris et Anderson, 1996*]. The sinistral fault propagator model includes this obliquity into the continuum of the exhumation of the MCCs.

A strike-slip boundary of the Aegean-west Anatolian extensional system may have started ~25 Ma ago [*Jolivet et al., 2012*]. Between 25 and 15 Ma, the total surface displacement of the strike-slip boundary can be measured by the amount of extension in the northern Menderes Massif resulting in some 50-60 km [*van Hinsbergen 2010*]. Until now, a lot of putative structures have been proposed to accommodate strike-slip displacement in the area with different orientation and kinematics [*Gautier and Brun 1994, Walcott and White,1998, Philipon et al. 2012*], however no large strike-slip structure has ever been clearly evidenced from field data or seismic reflection data. Our numerical experiments show that the Naxos structure is better explained as a transtensional structure rather than a purely extensional one. We argue that the shape of the dome of Naxos [*Vanderhaeghe, 2004*, *Kruckenberg et al., 2011*] was acquired as it plays the role of a lower crustal root of a strike-slip fault which was active in the Miocene times. Whether the a-type shape of the dome of Naxos marks a local segment jump between the Menderes massif and the Northern Cycladic Detachment System, or whether it is driven by global geodynamics is a matter of debate which we cannot quantitatively assess based on the results of our numerical experiments. However, to conclude on the Aegean dynamics, we would like to propose a completely qualitative and conceptual, yet deliberately provocative model for the formation of the splay fault that we imposed in the boundary condition of our preferred Aegean model.

A tear within the Aegean slab is documented by Berk-Biryol *et al.* [*2011*] under western Turkey and propagating along the Pliny and Strabo trenches south east of Crete and corresponds today to a steep lateral gradient in trench retreat rate. The timing of the initiation of this tear is debated. Pe-Piper et Piper [*2007*] and Jolivet et al. [*2009*] evaluate it at c.a. 20 Myr based on adakites ages (Fig. 8 a. and b.) while van Hinsbergen et al. [*2010*] posit this tear is not older than 15 Myr based on palinspatic plate reconstruction. Reconstruction of the Aegean area [*Philippon* *et al., 2012*] *show* that in the Miocene time, Anatolia and the Cyclades were probably further east than the tears current location. The strike-slip wrenching recorded in the a-type structure of Naxos could well be the lower crustal signature of a step fault accommodating the steep gradient in slab retreat rate caused by the slab tear (Fig. 8c). Within that hypothesis, the tear in the slab must initiate prior to 15 Myr, which corresponds to the late evolution of that dome [*Duchène et al., 2006*].

6 General discussion and conclusions

Previous models for the formation of non-cylindrical domes in the crust involved components of normal shortening [*Thompson et al., 1997*], normal stretching [*Rey et al., 2011*] or any complex poly-phased history which would include sequences of wrenching and folding or a late refolding of a pre-existing b-type dome with a shortening direction normal to the direction of extension [*Philipon et al 2012*]. All these equally valid models emphasize the non-uniqueness of the solution when trying to reconstruct the kinematics from finite-strain markers. Here, we show that the using a strike-slip fault propagator, or an extensional step-over in the boundary conditions of numerical experiments, together with initial conditions which favor the formation of MCCs, provides yet another valid alternative model for the emplacement of non-cylindrical domes. However, the main point we would like to emphasize is that the presence of constrictive ductile dome structures is not a sufficient argument to infer transpressive boundary conditions, or far field shortening in the absence of a geochronological argument for a more complex poly-phased deformation history, as was proposed in earlier models.

Comparing the numerical experiments with the Aegean, we propose an alternative model to the formation of the Cyclades and relate the N-S axis of the Naxos dome to a phase of transtension. A possible source for the transtension could be the presence of a tear in the Aegean slab, but other hypothesis such as a transfer fault between the Cyclades and the Menderes massif (Western Turkey) remain as a possibility. In any case, we argue that the elongation and the structure of the Naxos dome is an indicator that the dome acted as a strike-slip structure during its activity. An argument to support this model is that on the opposite side of the Mediterranean, a-type domes are clearly observed in the Betics [*Augier, 2005*] and are again aligned to a tear propagating in a slab [*Spackman and Wortel, 2004*].

The numerical experiments and conclusions drawn in this paper are directly applicable to MCCs other than those in the Aegean. However, given the availability of data from the Aegean, this region represents an excellent location to validate our ideas and verify our 3D thermo-mechanical models. Our numerical experiments constitute a first order study of the mode of 3D deformation of thickened crust. Each of these experiments would require a systematic parametric study varying rates of extension or shear, relative timing between the onset of shear and extension, depth of the viscosity step, length scale of the step over in order to be applied and validated with data in different geological settings and geodynamic contexts.

The aim of this paper was not to perform such parametric studies, but to open the door to alternative models and interpretations for the 3D ductile deformation of post-orogenic, relatively thickened crust. With our simulations, we have shown that it is possible to form folds in response to local constriction within extensional step-over between strike-slip faults (model 2, Figure 5). This model setup may well apply to the Red River Fault system along which many discontinuous exhumed domes are out-cropping [*Leloup et al., 1995*]. Within this model, these domes would be the mark of extensional step-over within a purely strike-slip system. Similarly, many of the Variscan gneiss domes which form *en echellon* are usually interpreted as transpressional structures (e.g. Pyrenees [*Denèle et al., 2007*]). Our model shows they could alternatively have formed below an extensional step over within a purely strike-slip, large-scale kinematic boundary conditions. The formation of these domes during the early stage of wrenching could also participate in weakening the strike-slip faults by advecting hot/weak material within the fault zone and locally reducing the effective elastic thickness of the fault zone as was proposed by [*Chery, 2008*] to explain the GPS signal around the San Andreas Fault.

Further work is needed to completely understand the dynamics of strike-slip systems in the ductile part of the crust, however as 3D thermo-mechanical modeling is now becoming readily available, we can expect much progress to be made in the near future.

APPENDIX

A Computation of synthetic tectoglyphs

The method used for the computation of the synthetic finite-strain field is an extension to 3D of a classical 2D method of Ramsay and Huber [*1987*]. The synthetic finite-strain field is computed over a set of passive markers, which is meshed with tetrahedra. The schistosity and lineation are computed for all tetrahedra in the following way.

Each tetrahedron is described by three non-colinear vectors that connect the four vertices. These vectors are noted

$$\mathbf{a}_i = \left(x_i, y_i, z_i\right)^T, \quad i = 1,2,3 \qquad \text{(A.1)}$$

in the initial (undeformed) configuration and

$$\mathbf{A}_i = \left(X_i, Y_i, Z_i\right)^T, \quad i = 1,2,3 \tag{A.2}$$

in the final (deformed) configuration, where the sign ^{T} designates the transpose operator. The two sets of vectors are gathered in two square matrices,

$$\mathbf{u} = \left(\mathbf{a}_1, \mathbf{a}_2, \mathbf{a}_3\right) \tag{A.3}$$

for the initial configuration and

$$\mathbf{U} = \left(\mathbf{A}_1, \mathbf{A}_2, \mathbf{A}_3\right) \tag{A.4}$$

for the final configuration.

The deformation matrix **D**, which relates the initial coordinates of the tetrahedron vectors to the final ones, links the matrices **u** and **U**:

$$\mathbf{U} = \mathbf{D}\mathbf{u} \quad \text{or} \quad \mathbf{D} = \mathbf{U}\mathbf{u}^{-1} \tag{A.5}$$

The finite strain ellipsoid is characterized by the eigenvalues $I_1 \leq I_2 \leq I_3$ and the corresponding eigenvectors $\mathbf{v}_1$, $\mathbf{v}_2$ and $\mathbf{v}_3$ of matrix

$$\mathbf{E} = \left(\mathbf{D}\mathbf{D}^{-1}\right)^{-1} \tag{A.6}$$

The major (resp. intermediate and minor) principal axis of the finite strain ellipsoid is parallel to $\mathbf{v}_1$ (resp. $\mathbf{v}_2$ and $\mathbf{v}_3$). The length of the major (resp. intermediate and minor) principal axis is $I_1^{1/2}$ (resp. $I_2^{1/2}$ and $I_3^{1/2}$). Finally, the stretching lineation is a line parallel to $\mathbf{v}_1$ and the schistosity is a plane normal to $\mathbf{v}_3$. All these computations can be done using the script *tectoglyphs.m* provided with the manuscript.

B Multi-grid in GALE

Within GALE, the discretized Stokes equations are represented as follows;

$$\begin{pmatrix} \mathbf{K} & \mathbf{G} \\ \mathbf{D} & \mathbf{C} \end{pmatrix} \begin{pmatrix} \mathbf{u} \\ \mathbf{p} \end{pmatrix} = \begin{pmatrix} \mathbf{f} \\ \mathbf{h} \end{pmatrix} \tag{B.1}$$

The velocity pressure ($\mathbf{u}, \mathbf{p}$) solution is obtained via Schur Complement Reduction. This entails applying a Krylov method to the Schur complement system

$$\left(\mathbf{D}\mathbf{K}^{-1}\mathbf{G} - \mathbf{C}\right) \mathbf{p} = \hat{\mathbf{f}}, \tag{B.2}$$

Where

$$\hat{\mathbf{f}} = \mathbf{D}\mathbf{K}^{-1}\mathbf{f} - \mathbf{h} \ , \tag{B.3}$$

to obtain the solution for $\mathbf{p}$. Following this, we apply another Krylov method to

$$\mathbf{K}\,\mathbf{u} = \mathbf{f} - \mathbf{G}\,\mathbf{p}\,, \tag{B.4}$$

in order to obtain the solution for $\mathbf{u}$. In GALE, the Schur complement system in (B.2) is solved using the Conjugate Gradient method. Iterations are terminated when the initial residual has been reduced by a factor of 1e5. Systems involving $\mathbf{K}$ are solved using FGMRES, using a stopping condition requiring that the initial residual be reduced by a factor of 1e6. The same Krylov method used to define the matrix-vector product in Eqn. (B.2) is used in both Eqn. (B.3) and Eqn. (B.4).

The performance of a Krylov method is strongly linked to the preconditioner used. The ideal preconditioner is both scalable and optimal; meaning that number of iterations required to reach convergence is constant as the finite element mesh is refined, and that the CPU time required to reach convergence scales linearly with the number of unknowns. For the system in Eqn. (B.2), GALE uses a preconditioner defined by the mass matrix, scaled by the inverse of the element-wise effective viscosity. For Q1-Q1 velocity-pressure spaces, this preconditioner has been proven [*Grinevich and Olshanskii, 2009*] and demonstrated for geodynamic applications [*Geenen et al., 2009; Burstedde C. et al., 2009*] to produce iteration counts for the Schur complement system that are independent of the mesh resolution.

To achieve an overall optimal solution strategy for the complete Stokes solver, GALE employs a geometric multi-grid (GMG) preconditioner for systems of the form $\mathbf{K}\,\mathbf{y} = \mathbf{x}$. A geometric multi-grid preconditioner utilizes a hierarchical representation of the discrete operator **K**. The simplest hierarchy contains two levels, consisting of **K** (fine level) and $\mathbf{K}_c$, a “coarse” grid operator which is defined on a coarser finite element mesh. In the classical multi-grid strategy, one removes high frequency components of the error in the solution by applying several iterations of an iterative method, such as Richardson+Jacobi or Richardson+Gauss-Seidel. Due to their properties, these methods are called "smoothers". Since only long wavelengths remain, the smoothed error can be represented on a coarser grid. The inter-grid transfer from the fine-to-coarse level is called "restriction". On the coarse grid, the exact error can be computed using a direct solver, as the coarse problem contains fewer degrees of freedom than the fine grid. The low-frequency errors on the coarse grid are transferred to the fine grid in order to correct the fine grid approximation for the low-frequency errors. The inter-grid transfer from the coarse-to-fine grid is called "prolongation". The smoother is also applied to the corrected fine grid approximation. We refer to Wesseling [*1992*], Briggs et al. [*2000*] and Trottenberg et al. [*2001*] for an in-depth discussion of multi-grid theory.

To obtain "text-book" efficiency multi-grid performance, the coarse grid operator has to be a meaningful approximation of the fine grid operator **K**. For partial differential equations (PDEs) containing coefficients which are highly spatially variable (continuous or discontinuous) and heterogeneous (e.g. viscosity), this is difficult within the framework of GMG. Primarily this is due to the inability of the coarse grid to adequately resolve fine scale structures. Additionally, in practice, the classical smoothers are found to be largely ineffective for problems which possess large variations in coefficients. We note that in the instance of discontinuous coefficients, if the coarse grid exactly resolves the coefficient jump, all the aforementioned issues are eliminated. However, for most practical geodynamic models employing markers - this scenario will rarely ever occur. In summary, the construction of the coarse grid operator and the choice of the smoother are extremely important for the development of efficient GMG preconditioners when they are applied to problems which exhibit large variations in coefficients.

In GALE, the restriction operation **R** is defined via bilinear interpolation between the fine and coarse grid levels. The prolongation operator **P** is defined as the transpose of **R**. Galerkin coarse grids operators are utilized on all levels (except the finest) in which the coarse grid operator is constructed via $\mathbf{K}_c = \mathbf{R}\mathbf{K}\mathbf{R}^T$. This type of coarse grid can be interpreted as a projection of the entire fine grid matrix onto the coarse grid. In general, practitioners often don't utilize Galerkin coarse grid operators due to the programming complexity (particularly in parallel) of defining the triple matrix product required. One alternative to Galerkin coarse grid operators is simply to project the coefficients onto the coarse grid and re-discretize the original PDE. However, in practical applications with large variations in coefficients, the Galerkin coarse grid operator produces a much more robust preconditioner. In addition, the "smoother" used in GALE is a fully-fledged Krylov method (FGMRES) equipped with a block Jacobi+ILU preconditioner. On both the down and upward sweep in the multi-grid V-cycle, we apply 8 iterations of this Krylov method. In contrast to the classical Richardson+(Jacoboi/Gauss Seidel/SOR) type smoothers, our practical experience of using such "heavy" smoothers have proved to be robust when applied to problems involving large variations in viscosity. Furthermore, on the coarsest grid level, we utilize a fully parallel LU factorization SuperLU_Dist [*Xioye, 2005*]. All linear algebra, Krylov methods and preconditioners are provided via PETSc v3.0 [*Balay et al., 2008*].

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TABLES

Table 1: ***Model parameters***

Notation/Unit	Lower crust	Upper crust	Mantle
η_0 (Pa.s)	10^{26}	10^{26}	10^{26}
θ (°C^{-1})	0.035	0.020	0.012
η_{eff} at 300°C (Pa.s)	$3x10^{21}$	$2x10^{23}$	$3x10^{24}$
η_{eff} at 700°C (Pa.s)	10^{19}	$8x10^{19}$	$2x10^{22}$
η_{eff} at 1100°C (Pa.s)	10^{19}	10^{19}	$2x10^{19}$
Co (Pa)	$2x10^{7}$	$2x10^{7}$	$2x10^{3}$
Co_{inf} (Pa)	$2x10^{7}$	$2x10^{7}$	$2x10^{3}$
φ (°)	30	30	15
φ_{inf} (°)	10	10	15
α (K^{-1})	$3x10^{-5}$	$3x10^{-5}$	$3x10^{-5}$
χ (m^{2}s^{-1})	10^{-6}	10^{-6}	10^{-6}
ρ (kg m^{-3})	2800	2800	3300

Table 2: ***Scaling used for computations***

Notation/Unit	Model	Nature (SI)	Scaling (SI)
Velocity $[LT^{-1}]$	10^5	10^{-10}	10^{-15}
Distance [L]	1	10^5	10^5
Time [T]	10^{-6}	10^{14}	10^{20}
Viscosity $[ML^{-1}T^{-1}]$	10^{-4}	10^{19}	10^{23}
Stress $[ML^{-1}T^{-2}]$	10^3	10^6	10^3
Density $[ML^{-3}]$	1	10^3	10^3
Gravity $[LT^{-2}]$	10^6	10	10^{-5}
Thermal diffusivity $[L^2T^{-1}]$	10^4	10^{-6}	10^{-10}
Temperature [°C]	1	1	1

FIGURE CAPTIONS

Figure 1: On the right, a 3D sketch (modified from Jolivet et al. [*2004*]) illustrates two classes of domes. On the left, a sketch illustrates the stereo-plot projections of the lineation (L) and foliation (S) for the two kinds of dome. In the a-type dome, constriction is important and the foliation is folded with axis aligned with the direction of stretching. In the b-type dome, the foliation is folded with an axis normal to the direction of extension. The a-type domes are not cylindrical and cannot be modeled in 2D.

Figure 2 : a. Initial geotherm is linear with a thermal gradient of 17°C/km down to the 1300°C isotherm and constant deeper; b. Lithostratigraphic column with density u.c., l.c. and m. stands for upper crust, lower crust and mantle respectively; c. Yield strength envelopes drawn in extension for a background strain rate equal to the mean strain rate (black line, $1.5\times10^{-15}\ s^{-1}$) and ten times the mean strain rate (red line, $1.5\times10^{-14}\ s^{-1}$).

Figure 3: Kinematic boundary conditions, surface deformation and topography with equidistance of 100 m for the three models after 12 Myr evolution. The black, gray, cyan and pink line outline respectively the right, left, back and front face of the model. Brown lines initially formed squares. a. Model 1 assumes cylindrical extension; b. Model 2 corresponds to a 60 km wide extensional step-over; c. Model 3 represents a transtensional fault propagator.

Figure 4: Results for cylindrical extension (Model 1). a. Internal deformation of the model outlined by cross-sections across the material points and by tubes representing the stretching lineation (maximum stretching axis of the finite strain tensor). The tubes are colored by their strike with color scale represented in b. where gray indicates when the

lineation is aligned with the direction of stretching imposed at the boundary of the model; b. Stereo-plot representation of the lineation (red) and the foliation (black) for all the tracers located at less than 8 km depth after 12 Myr of simulation; c. Synthetic P-T path for the same tracers as those represented in the stereo-plot in b. Initial and final thermal gradient in blue and yellow respectively. The final thermal gradient is constrained assuming the line goes though 0° at the surface.

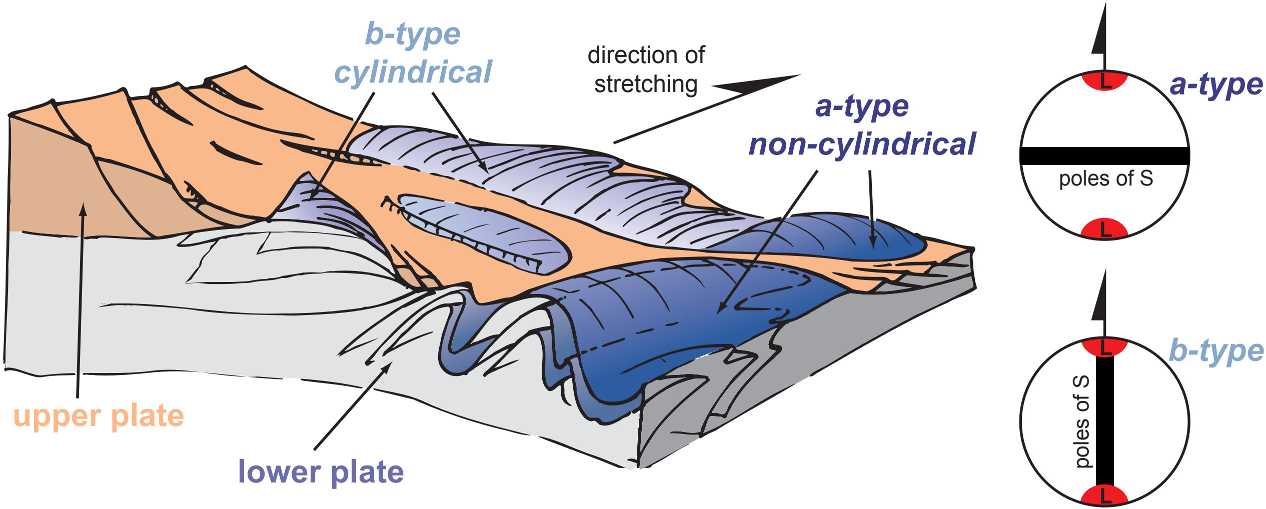
Figure 5: Results for extensional step over (Model 2), legend is the same as for Figure 4

Figure 6: Results for transtensional fault propagator (Model 3), legend is the same as for Figure 4

Figure 7: a. Topographic map of the Cyclades with outline of the Cycladic MCCs (purple area), as well as stretching lineations and sense of shear (arrows drawn after Martin [2004], Huet et al. [*2009*], Iglseder et al. [*2009*], Jolivet et al. [*2010*] and Ring et al. [*2011*]). The different colors outline different provinces associated with typical strikes using the same color scale as in Fig 4. The modeled area is reported with white dashed lines, black boxes denotes where the cylindrical part and non-cylindrical part were sampled in the model 3 to produce the part c and d of the figure; b: Geological sketch representing the main structural features of model 3; c: The blue and magenta lines are the natural P-T paths of Tinos [*Parra et al., 2002*] and Naxos [*Duchène et al., 2006*] respectively. Light and dark gray lines denote paths located within non-cylindrical and cylindrical part of model 3 (see a); d: Stereo-plot of the stretching lineation (red) and pole of the foliation (black) in the cylindrical (left) and non-cylindrical (right) part of model 3 oriented in the same frame as the map in a. The cylindrical part is a typical b-type dome while the non-cylindrical part

resembles an a-type dome with a non negligible number of lineation oriented N-S like in Naxos.

Figure 8: a and b Paleogeographic reconstructions of the Cyclades and slab tear simplified from [*Jolivet et al., 2009*], adakite from [*Pe-Piper and Piper, 2007*]; c. Sketch indicating how a slab tear could be responsible for the strike component imposed at the boundary of model 3.



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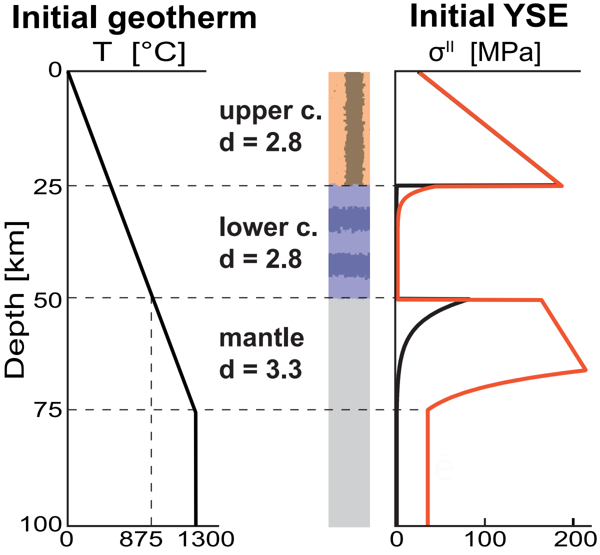

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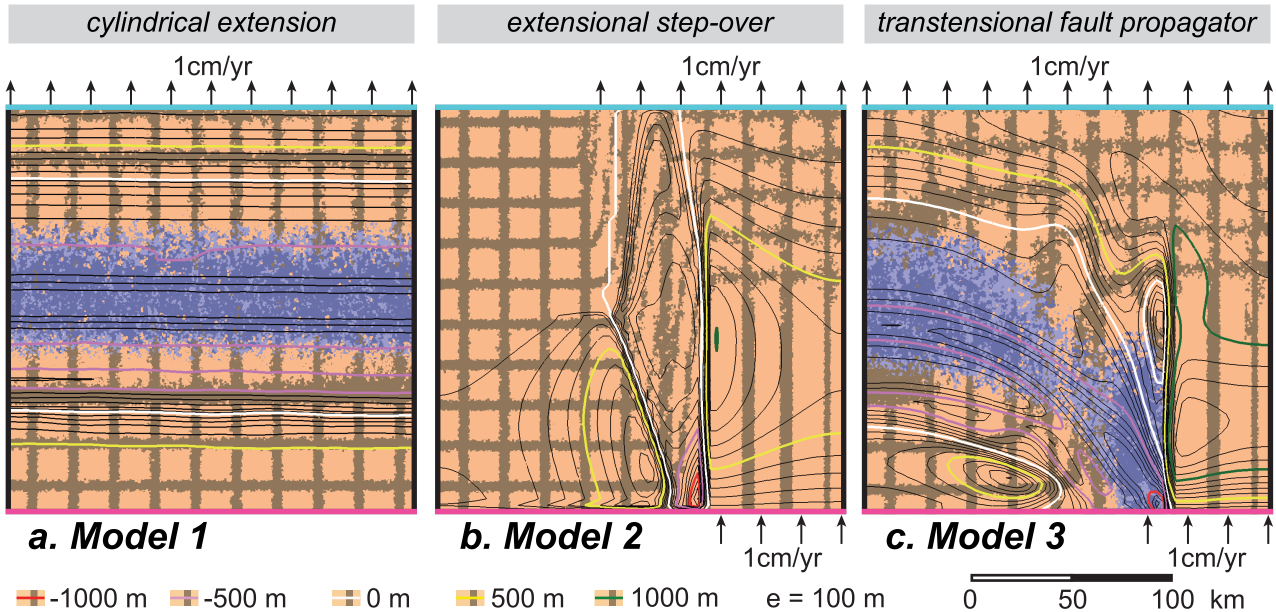

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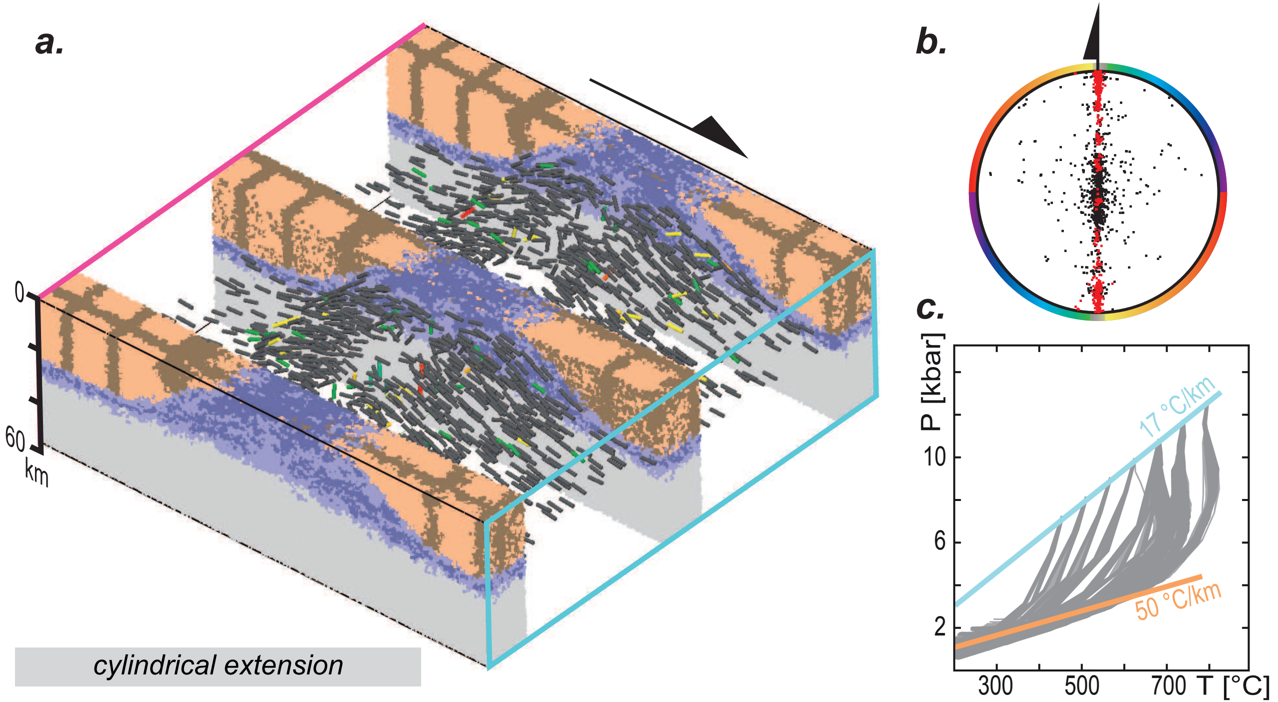
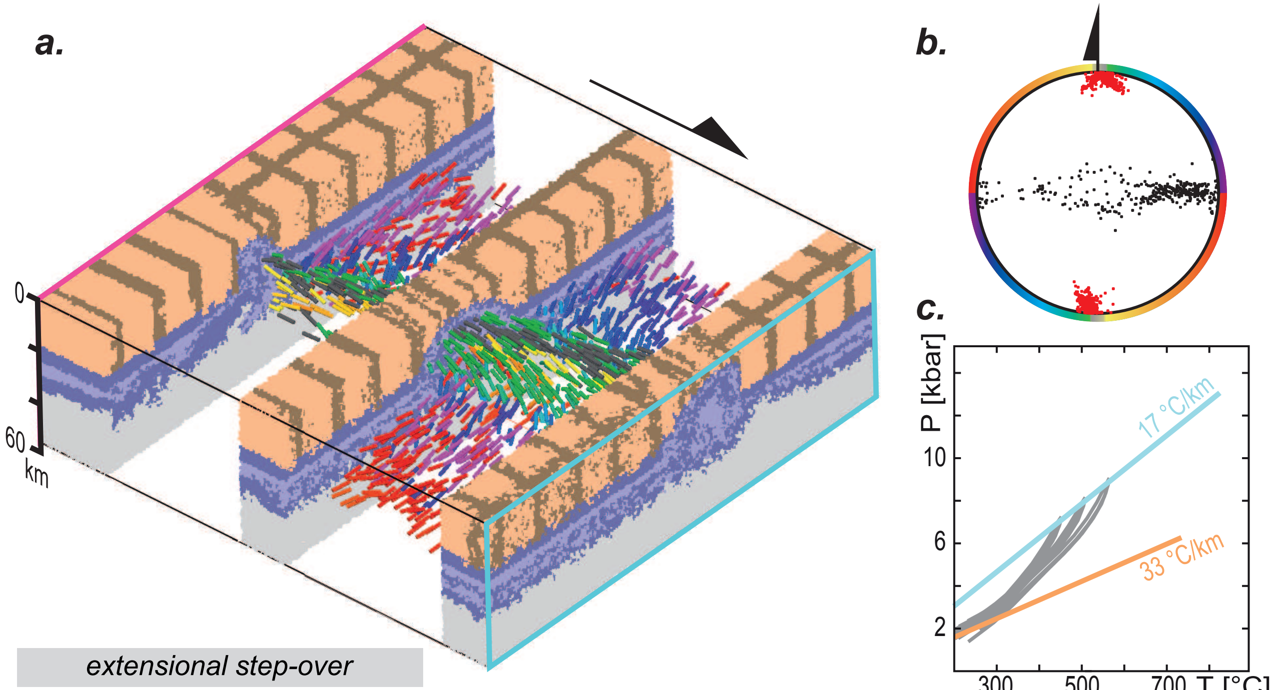


Figure 4: Results for cylindrical extension, Model 2, a. internal deformation of the model outlined by cross-sections across the material points and by tubes representing the stretching lineation (maximum stretching axis of the finite strain tensor). The tubes are colored by their strike with color scale represented in b. gray being used when the lineation is aligned with the direction of stretching imposed at the boundary of the model; b. Stereo-plot representation of the lineation (red) and the foliation (black) for all the tracers located at less than 8km depth after 12 Myr of simulation; c. Synthetic PT path for the same tracers as those represented in the stereo-plot in b Initial and final thermal gradient in blue and yellow respectively. The final thermal gradient is constrained assuming the line goes though 0° at the surface.



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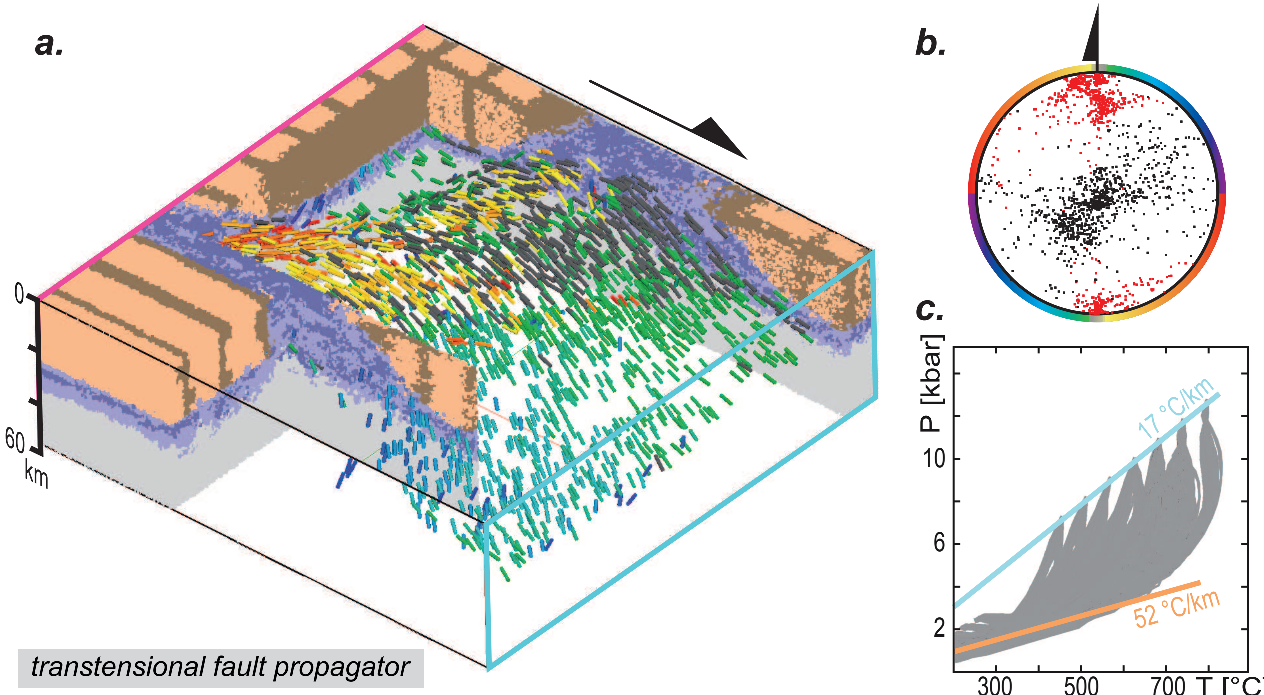
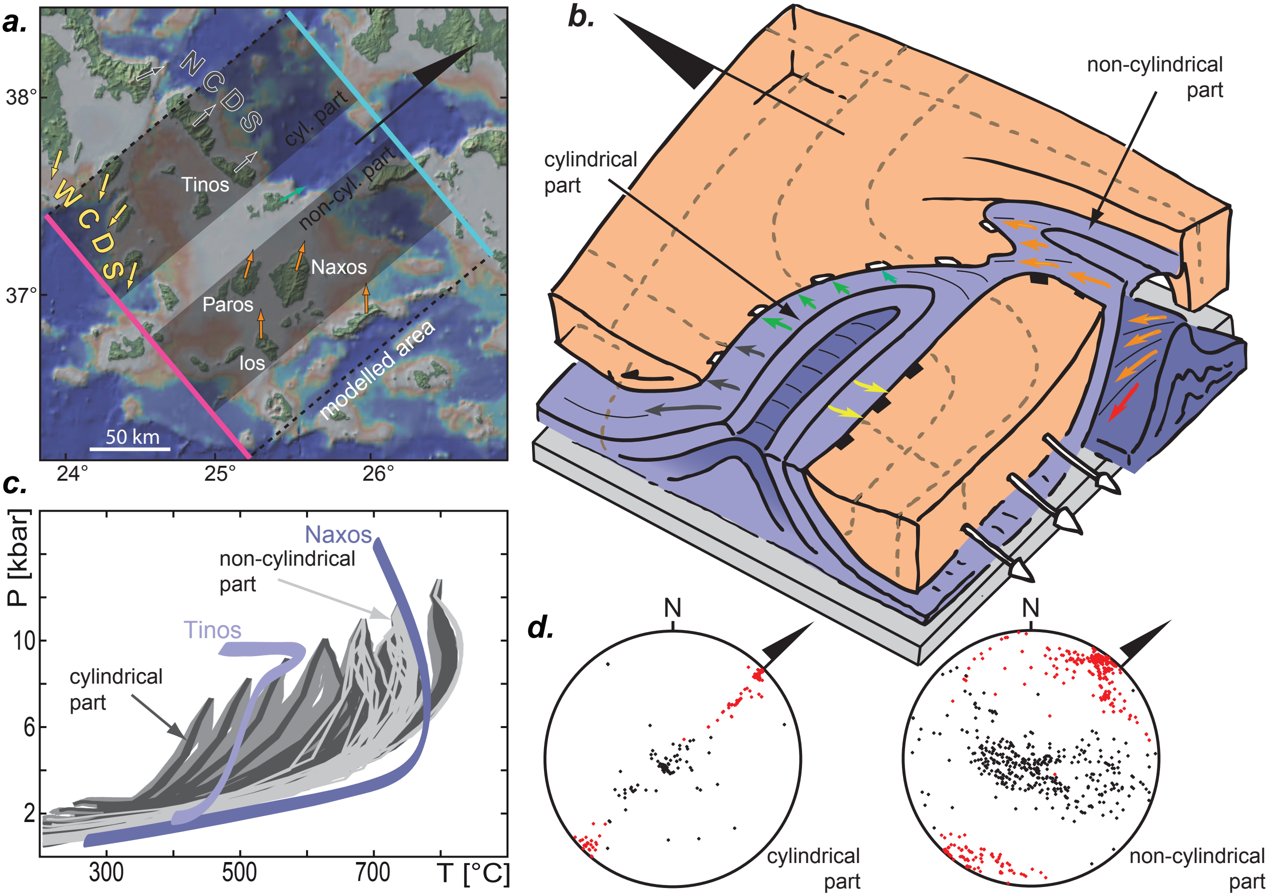


Figure 6: Results for trans-tensional fault propagator, Model 3, legend is the same as for Figure 4



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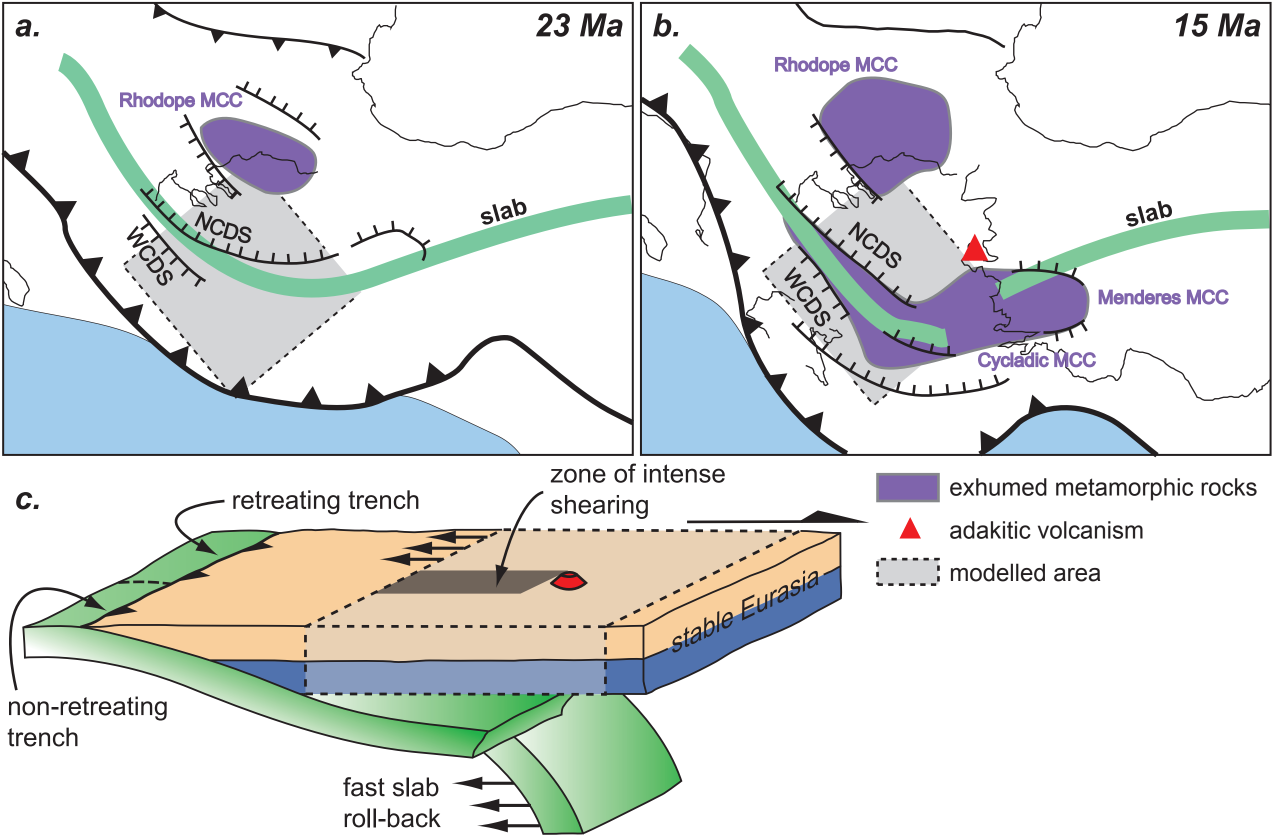

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